Final Laurentide ice-sheet deglaciation and Holocene climate-sea level change

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ABSTRACT
Despite elevated summer insolation forcing during the early Holocene, global ice sheets retained nearly half of their volume from the Last Glacial Maximum, as indicated by deglacial records of global mean sea level (GMSL). Partitioning the GMSL rise among potential sources requires accurate dating of ice-sheet extent to estimate ice-sheet volume. Here, we date the final retreat of the Laurentide Ice Sheet with 10Be surface exposure ages for the Labrador Dome, the largest of the remnant Laurentide ice domes during the Holocene. We show that the Labrador Dome deposited moraines during North Atlantic cold events at ~10.3 ka, 9.3 ka and 8.2 ka, suggesting that these regional climate events helped stabilize the retreating Labrador Dome in the early Holocene. After Hudson Bay became seasonally ice free at ~8.2 ka, the majority of Laurentide ice-sheet melted abruptly within a few centuries. We demonstrate through high-resolution regional climate model simulations that the thermal properties of a seasonally ice-free Hudson Bay would have increased Laurentide ice-sheet ablation and thus contributed to the subsequent rapid Labrador Dome retreat. Finally, our new 10Be chronology indicates full Laurentide ice-sheet had completely deglaciated by 6.7 ± 0.4 ka, which requires that Antarctic ice sheets contributed 3.6 – 6.5 m to GMSL rise since 6.3 – 7.1 ka.

1. Introduction
During the early to middle Holocene (11.7–6.0 ka), warming from boreal summer insolation and greenhouse gases (Marcott et al., 2013) caused the disappearance of most Northern Hemisphere ice sheets. Despite this strong radiative and temperature forcing, global mean sea level (GMSL) was still ~60 m below present at the start of the Holocene (Lambeck et al., 2014), identifying a pronounced lag between climate forcing and subsequent ice-sheet mass loss. Partitioning this GMSL rise into its individual ice-sheet contributions remains uncertain, thus preventing a full assessment of ice-sheet sensitivity to climate change that is critical to understanding glacial-interglacial cycles (Bintanja and van de Wal, 2008; Carlson and Winsor, 2012; Tarasov et al., 2012; Abe-Ouchi et al., 2013; Ullman et al., 2015a, 2015b; Stokes et al., 2016).

The Laurentide Ice Sheet (LIS) was the largest contributor to Holocene GMSL rise (Peltier, 2004), but its retreat history is poorly constrained by minimum-limiting 14C dates that may underestimate the age of deglaciation by up to several thousand years (Dyke, 2004; Carlson et al., 2007, 2008). Better constraints on the LIS retreat history would improve our understanding of its contribution to Holocene GMSL rise as well as address the question of its sensitivity to climate forcing, including whether its retreat was...
modulated by centennial climate variability that may itself have originated from forcing by the LIS (Hillaire-Marcel et al., 1981; Alley et al., 1997; Bond et al., 1997; Barber et al., 1999; Clark et al., 2001; Kleiven et al., 2008; Yu et al., 2010; Carlson et al., 2009b; Hoffman et al., 2012; Carlson and Clark, 2012; Jennings et al., 2015).

Here we focus on the final deglaciation of the Labrador Dome (LD) (Fig. 1), the largest of the LIS domes (Fig. 1) (Dyke, 2004). We report $^{10}$Be surface exposure ages measured on glacial erratic boulders along 3 transects in Quebec and Labrador to directly date deglaciation, circumventing issues with minimum-limiting $^{14}$C ages (Porter and Carson, 1971). In addition, we present results from new high resolution regional climate model simulations that suggest the loss of ice over Hudson Bay provided an additional forcing driving the negative mass balance of the LD. Finally, we use this new chronology to estimate the LD contribution to GMSL rise across the Holocene allowing us to infer the remaining contribution from Antarctic ice sheets.

2. Methods

2.1. Geomorphic setting and field methods

The sampling area is typical of the glacially-smoothed sub-arctic Canadian Shield. Boreal spruce forest and glacially-carved lake basins cover much of this low relief landscape, with exposed topographic highs dominated by streamlined and striated outcrops.
of exposed Precambrian igneous and high-grade metamorphic bedrock with minimal till cover. Large glacial erratic boulders of granitic or gneissic lithology are prevalent. Decades of extensive logging and hydroelectric power generation has affected much of the region, but numerous bedrock topographic highs do not support forest cover, resulting in parts of the landscape that have not been altered since deglaciation.

We attempted to limit the impact of human disturbance and post-depositional movement in exposure surfaces by sampling the surfaces of large (>1 m in diameter) quartz-bearing glacial erratic boulders (Fig. 2) on stable bedrock topographic highs with minimal till cover. Isostatic depression led to the incursion of marginal seas along the coasts, so near-coastal sites (QCNS and CL1) were selected to be above the regional marine limit (Andrews and Peltier, 1989) to avoid the impacts of post-glacial submergence of the boulders (Carlson et al., 2007).

Samples were collected using a hammer and chisel from the tops of boulders. Surfaces with evidence of erosion (pitting or spallation) were avoided; we therefore assume a zero erosion rate for these samples. Geographic location, elevation, and thickness were recorded (Supplementary Table 1). Topographic shielding was not an issue due to the selection of topographic highs as sampling sites.

We report 65 new boulder samples along 2 transects perpendicular to LD margin retreat and combined these with recalibrated existing 10Be ages (Carlson et al., 2007) along a 3rd transect (Fig. 1). Along this 3rd transect, we also reanalyzed 3 of the original samples and 1 additional sample (QC-26) not included in Carlson et al. (2007).

2.2. Sample processing and analytical techniques

Sample preparation and beryllium oxide isolation was conducted in the Cosmogenic Nuclide Laboratory at the University of Wisconsin-Madison. Samples were crushed and sieved to separate the 425–841 μm grain-size fraction. A Frantz magnetic separator was used to remove magnetic grains, followed by repeated etchings with HCl and dilute HF/HNO3 to isolate the quartz fraction. The purity of this quartz fraction was measured through elemental analysis by ICP-OES at the University of Colorado-Boulder.

Samples were converted to BeO through a series of dissolution, oxidation, cation/anion removal, pH adjustment steps, and ignition. A Be standard of known concentration (Murray et al., 2012) was added to each sample to facilitate calculation of 10Be concentrations from the 10Be/9Be ratios measured by accelerator mass spectrometry at PRIME Laboratory at Purdue University (Supplementary Table 1). With each batch of samples, a procedural blank was processed to measure the ambient levels of contamination in the cosmogenic lab that accumulate throughout BeO isolation; average blank 10Be/9Be ratios were 2.0 ± 0.34 × 10⁻¹⁵ (n = 18).

2.3. Exposure age calculation

10Be ages were calculated using the CRONUS-Earth online calculator (v 2.2) (Balco et al., 2008) using the northeastern North America regional production (Balco et al., 2009). In our analysis and presentation of results, we use the time-varying Lal (1991) and Stone (2000) scaling scheme (Balco et al., 2008). Use of other neutron-monitor scaling schemes (Dunai, 2001; Lifton et al., 2005; Desilets et al., 2006) does not alter our results (see Supplementary Table 2) or interpretations, as the differences (<5%) lie within the analytical uncertainty of each measurement.

2.4. Post-glacial uplift and atmospheric pressure corrections to 10Be production rate

Northeastern Canada has experienced significant post-glacial rebound since the initial exposure of our sites, as indicated by the fall of local sea level (Andrews and Peltier, 1989). The effect of this uplift has the possibility to significantly impact 10Be production rates since deglaciation, slowly increasing through time. Young et al. (2013) argued that the effect of post-glacial rebound on 10Be production rate is offset “to an unknown degree” by changes in atmospheric pressure due to local air pressure evolution in a changing deglacial climate (Stone, 2000; Staiger et al., 2007), which we test below.

Following Cuzzone et al. (2016), we quantify the time-varying effects of uplift and atmospheric pressure on the 10Be production rates since exposure of our sites. To estimate site-specific altitude changes since deglaciation, we apply an isotropic surface loading model (Mitrovica et al., 1994) that includes the influence of ice loading (using the ICE-5G reconstruction of ice thickness and its partnering Earth viscosity model, VM2; Peltier, 2004), ocean loading (Mitrovica and Milne, 2003) and variations in Earth rotation (Mitrovica et al., 2005). This approach explicitly estimates the true vertical land motion, without the confounding effects of GMSL rise and the gravitational attraction of the remaining ice sheets, which are implicitly included in any estimate from local relative sea level data. Uplift corrections range from 20 to 105 m. These corrections result in 10Be ages that are 1–10% older than non-uplift-corrected ages, with spatial variability due to the magnitude of post-glacial rebound. Modern and corrected elevations for each sample are provided in Supplementary Table 1.

To estimate the impact of evolving atmospheric pressure due to changing post-glacial climate (Stone, 2000; Staiger et al., 2007), we use output from a coupled atmospheric-ocean general circulation model providing simulated climate at 3 ka time slices from 21 ka to 0 ka (Alder and Hostetler, 2015). After linearly interpolating this...
output between time slices, we estimate the change in atmospheric thickness due to the change in surface air pressure using the hypsometric equation with output from these deglacial time slice climate simulations (Cuzzone et al., 2016). Through this method, we determine the average elevation correction for our sites range from 3 m to 6 m (or 1–5% of site elevation). While opposing the uplift correction in direction, this small atmospheric correction does not offset the uplift correction in magnitude. In addition, the atmospheric correction is within the accuracy of our site measurement of elevation. Therefore, we exclude the atmospheric correction in our overall topographic correction and only account for isostatic rebound in our age estimates.

2.5. Averaging samples

We use Chauvenet’s criterion to identify outliers that have a probability of <50% of falling within the normal distribution of the sample set (Rinterknecht et al., 2006). This method first requires a normally distributed dataset, which we confirm for all sample groupings with n > 2 using a Shapiro-Wilk test for normality; each group of n = 2 overlap within 1-sigma analytical uncertainty. As a visual test of normality shown by the Shapiro-Wilk test, we present our data in a quantile-quantile plot for each sample grouping with n > 2 (Fig. 3).

To determine the most appropriate method of averaging for each site, we compare the geological uncertainty of the sample group to the analytical uncertainty (e.g., Douglass et al., 2006; Rinterknecht et al., 2006; Carlson et al., 2007; Murray et al., 2012; Ullman et al., 2015b). The relative magnitude of analytical and geological uncertainty is calculated using the mean square weighted deviation (MSWD) of each sample grouping (Supplementary Table 3), and we use this as a statistical means of determining when a sampling grouping should be averaged using a straight mean (with standard error as uncertainty) or an error-weighted mean (with error-weighted sigma as uncertainty). If MSWD is > 1, then the scatter in the ages (geologic uncertainty) is larger than the analytical uncertainty and we therefore apply the straight mean approach of averaging. However, if MSWD is < 1, then the analytical uncertainty is larger than the scatter of the individual ages and we apply an error-weighted mean.

While we use MSWD as systematic tool for selecting most appropriate averaging method, we note that use of the alternate averaging method (i.e., using straight mean instead of error-weighted mean or vice-versa) at each site results in an average

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Fig. 3. Exposure age versus theoretical quantile for each of the sample groupings. This modified version of a Quantile–Quantile plot includes the internal uncertainty for each of the dates. The black line on each plot shows the expected result for a normal distribution (scaled with the mean and standard deviation of each sample grouping). Note: This plot is not possible for site CL2 because the sample size is 2.
that is consistent within uncertainty and does not significantly change our results or our interpretation. We also note that our approach of selecting the averaging method based on the relative magnitude of analytical versus geological uncertainties is the statistically conservative approach, where we choose the method with the greatest uncertainty. We finally include a constant production rate uncertainty of 4.8% (Balco et al., 2009), added in quadrature to the uncertainty of the mean.

2.6. Constructing time-distance transects

Assessment of ice-margin retreat rates using $^{10}$Be chronology requires a measure of distance between sampling sites. However, the direction of ice-margin retreat may not necessarily have followed the arrangement of our sample locations. Therefore, we draw time-distance transects perpendicular to the retreating ice margins of Dyke (2004) and parallel to the ice-flow patterns of Veillette et al. (1999). If a sample site lies off this transect, we project it perpendicularly (following ice margins and ice flow patterns) back to the transect. In addition, we project our mean age from the land-terminating portion of the Sakami Moraine (site SK; Fig. 1) onto the western transect by tracing the moraine from Lake Mistassini in southwestern Quebec to the beginning of our western transect (Hillaire-Marcel et al., 1981; Occhietti et al., 2011).

2.7. Estimating Laurentide ice-sheet volume

For regularly shaped ice caps/sheets, like the LIS, LD and other LIS domes, that approach a circular form, a maximum-estimate of ice volume can be approximated from areal extent (Paterson, 1994). This equation was determined empirically using a regression from data from Antarctica, Greenland, and other small ice caps of varying size and shape, and it has an uncertainty of ±12% (Paterson, 1994). We have updated this regression using newer estimates of ice sheet/cap volume and area than those used in Paterson (1994) (Fig. 4). This updated regression results in the empirical relationship expressed in the following equation:

$$\log V = 1.25 \log A - 1.13$$

where V is ice cap volume in km$^3$ and A is the areal extent in km$^2$. We use this new empirical relationship to estimate changes in LIS and LD volumes during the early to middle Holocene. Use of the Paterson (1994) relationship does not lead to different results, within uncertainty. We note that this relationship may be heavily influenced by the Greenland and Antarctic data points but their size is of similar order of magnitude to the LD.

2.8. Regional climate and surface mass balance modeling

In addition to the measurement of $^{10}$Be exposure ages, we conducted a pair of regional climate model simulations to test the impact of changing deglacial boundary conditions on the surface mass balance of the LD regional climate. Our model features a LIS topography that are important to the surface energy balance, including the ice-surface slope and the location of the ice margin relative to the shoreline of Hudson Bay (Fig. 5A), as well as regional atmospheric circulation and pressure (Fig. 5B).

Our modeling hierarchy is designed as follows. To establish global climate for ~8.5 ka, we added the LIS topographic reconstruction of Licciardi et al. (1998) to the 6 ka paleogeography of GENMOM, our coupled atmosphere-ocean general circulation model (AOGCM). GENMOM has a nominal atmospheric and oceanic resolution of 3.75° in latitude by longitude and includes the LSX surface physics model (Alder et al., 2011). Use of the 6 ka land mask includes the expanded area Hudson Bay that existed at ~8.5 ka as a result of isostatic depression. We prescribed 9 ka insolation and greenhouse gas forcings (Alder and Hostetler, 2015) and ran the simulation for 500 years to ensure that model spin up and stabilization of deep-ocean temperatures. GENMOM does reasonably well at capturing mid-Holocene temperature and precipitation as indicated by proxy data (Alder and Hostetler, 2015).

We derived a monthly sea-surface temperature (SST) climatology from the last 100 years of the GENMOM simulation and used the SSTs as a boundary condition in higher resolution version of GENESIS (3.75° × 3.75° atmosphere and 2° × 2° surface) in which we included the same configuration of the LIS and values of insolation and greenhouse gases. In the “ice” experiment we specified perennial sea ice over Hudson Bay and in the “no-ice” experiment we used the seasonal cycles of sea ice and surface temperatures from GENMOM. These scenarios simulate atmospheric conditions before/after the deglaciation of Hudson Bay based on what the atmosphere would have “felt” in terms of sensible and latent heat due to differences in ice cover. We ran the GENESIS simulations for 50 years and used the last 25 years to derive lateral (6 h vertical profiles of wind, temperature, and humidity) and surface (SST)
boundary conditions for the nested RegCM3 regional model (Pal et al., 2007) with a model resolution of 25 km and a setup that includes the BATS surface physics package (Dickinson et al., 1993). Our version of RegCM3 for paleoclimate applications includes the orbital parameter code from National Center for Atmospheric Research Community Climate System Model, which allows us to vary time-dependent insolation. We ran the regional model for 25 years and used the last 20 years in our analysis.

BATS uses a 3-layer soil model to account for surface hydrology as determined by precipitation, snow accumulation and melt, surface runoff and percolation. The surface temperature is determined by a force-restore method. The surface mass balance at a grid point \((i,j)\) is given by:

\[
M_{ij} = P_{ij} - E_{ij} - f_{ij}(R_{oi} + R_{ii})
\]

where \(M\) is mass balance, \(P\) is precipitation, \(E\) is evaporation, \(f\) is the fraction of snow melt and precipitation that runs off the ice, \(R_{oi}\) is runoff from precipitation and snow melt and \(R_{ii}\) is runoff from melted ice. We used the parameterization of Janssens and Huybrechts (2000) to compute the fraction of total runoff using monthly values of \(T\) and \(P\) from the regional model:

\[
f_{ij} = \begin{cases} 
0, & T_{ij} \leq 273.15 \\
1.0 - \frac{2c/Lf(273.15 - T_{ij})}{P_{ij} \times \text{ndays}}, & T_{ij} > 273.15 
\end{cases}
\]

where \(c\) is the specific heat of ice \((2090 \text{ J} \text{ kg}^{-1} \text{C}^{-1})\), \(L_f\) is the latent heat of freezing \((3.34 \times 10^5 \text{ J} \text{ kg}^{-1})\) and \(\text{ndays}\) is the number of days in the month. The fraction ranges from 0 to 1.

Similar to other climate models, BATS does not account for melting ice from ice sheets so when the seasonal snow accumulation is melted, runoff is generated only from precipitation. To capture ice melt, we added an additional term to the runoff equation that is based on widely used degree-day models:

\[
R_{ii} = \begin{cases} 
SWE \times (T_{\text{air}} - 273.15), & T_{\text{air}} > 273.15 \text{ and SWE} < 1\text{mm} \\
0, & T_{\text{air}} \leq 273.15 \text{ or SWE} \geq 1\text{mm} 
\end{cases}
\]

where \(k_{\text{ice}}\) is the degree-day coefficient \((10 \text{ mm m}^{-1} \text{K}^{-1})\) and SWE is the liquid water equivalent of the snow. In our model design, the degree-day melt is only applied to bare ice and is therefore decoupled from BATS surface model, so there is no conflict with the BATS calculated surface hydrology, which only calculates the snow melt. Use of a different degree-day factor would change the overall melt rate in the "no-ice" simulation, but the magnitude of melt anomalies between "no-ice" and "ice" (Fig. 5G) are independent of degree-day factor (so long as the same degree day factor is used for both simulations).

3. Cosmogenic surface exposure age results

Along the southern transect, sites QCNS and QCLW are on the North Shore Moraine, showing a joint deglaciation age of 9.2 ± 0.5 ka \((n = 10)\) as the average of the two sites (after excluding an old outlier from each site) (Supplementary Table 3). Farther north, CQ1 and CQ2 have mean ages of 8.2 ± 0.4 ka \((n = 7)\) and 7.7 ± 0.4 ka \((n = 6)\), respectively. The northernmost site on the southern transect, LC, has a mean \(^{10}\text{Be}\) age of 6.7 ± 0.4 ka. \(^{10}\text{Be}\) ages along this transect suggest average retreat rates of 70–1000 m a\(^{-1}\) with possible accelerated retreat ~8.2–7.1 ka (Fig. 6C). The overlap in the \(^{10}\text{Be}\) age uncertainties cannot rule out the possibility of near-instantaneous retreat from top-down melting (Carlson et al., 2007) occurring at some point during this time frame (Fig. 6C).
Comparison with $^{14}$C ages within 100 km of the southern transect shows an approximate centuries-scale lag behind our $^{10}$Be ages (Fig. 7B). (All $^{14}$C chronologies as discussed in the text have been recalibrated using Calib 7.1 and IntCal13; Stuiver and Reimer, 1993; Reimer et al., 2013).

At the start of our eastern transect, we average together CL1B and CL1 samples that are from the Paradise Moraine for a mean age of $10.4 \pm 0.6$ ka ($n = 10$; after excluding 4 outliers from CL1B) (Supplementary Table 3). CL2 has an average of $8.6 \pm 0.6$ ka ($n = 2$). CL3 and CL4 were deglaciated at $7.6 \pm 0.6$ ka ($n = 3, 2$ outliers) and $7.5 \pm 0.4$ ka ($n = 6, 1$ outlier), respectively. Our site LC that is the convergence of the southern and eastern transects was ice free at $6.7 \pm 0.4$ ka ($n = 5, 1$ outlier). Retreat rates averaged $40-310$ m a$^{-1}$ along the eastern transect, with a possible acceleration to near instantaneous retreat $-8.1-7.8$ ka (Fig. 6D). Minimum-limiting $^{14}$C ages along this transect lag our $^{10}$Be ages by $\sim 1$ ka (Fig. 7C).

Our western transect begins at the Sakami Moraine that ice retreated from at $8.2 \pm 0.5$ ka ($n = 8$). The seven remaining ages along the transect are consistent with rapid retreat to a cluster of ages in the east with deglaciation at $7.6 \pm 0.6$ ka ($n = 4$). Accordingly, rapid retreat averaged $860$ m a$^{-1}$ for the entire $520$-km transect (Fig. 6E), with the possibility of near instantaneous retreat $-8.2-7.6$ ka. The oldest $^{14}$C ages lag our $^{10}$Be ages by $-1$ ka along this transect (Fig. 7A).
4. Implications for climate and sea level

4.1. Moraine deposition within the context of North Atlantic climate variability

Our new \(^{10}\)Be ages for the Paradise (10.4 ± 0.6 ka), North Shore (9.2 ± 0.5 ka), and Sakami (8.2 ± 0.5 ka) moraines (Fig. 7C–E) suggest that, within uncertainty, the LD ice-margin stabilized at the same time as North Atlantic cold events occurring at ~10.3 ka, ~9.3 ka, and ~8.2 ka, respectively (Alley et al., 1997; Bond et al., 1997; Rasmussen et al., 2006) (Fig. 6A, B). Prior studies found west Greenland and Baffin Island outlet- and valley-glacier responses to the 9.3 and 8.2 ka events (Young et al., 2011, 2012), but our records are the first to show a response of the much larger LIS and the first evidence of a cryosphere response to the 10.3 ka event.

Detrimental carbonate deposition in the western Labrador Sea, along with geochemical records, provides a proxy of meltwater discharge events from Hudson Strait that may have forced these cold events (Carlson et al., 2009b; Hoffman et al., 2012; Jennings et al., 2015) (Figs. 1 and 6G). Climate model simulations show that Hudson Strait meltwater that enters the Labrador Sea can cause regional cooling of as much as 1.5 °C over the LD (Morrill et al., 2014), supporting our inference of a LD response to these early Holocene cold events. The climate model simulations also show minimal cooling (<0.5 °C) along the LIS margin west of Hudson Bay, which may explain why early Holocene moraines comparable to those of the LD are lacking (Dyke, 2004). We therefore suggest that these meltwater discharge events out of Hudson Strait and their attendant effect on climate may have helped sustain the LD while the western LIS margin continued to retreat rapidly (Dyke, 2004; Carlson et al., 2008, 2009a; Ullman et al., 2015a).

4.2. Final Laurentide ice-sheet deglaciation

Our data show that following the opening of Hudson Bay at ~8.2 ka (Barber et al., 1999), LD-margin retreat accelerated, although with spatial variability. Along our southern and eastern transects, LD margin retreat rates were up to orders of magnitude faster than those documented from the fjords of Baffin Island during the early Holocene (Briner et al., 2009). The overlapping ages from the western transect suggest century-scale loss of this sector between 8.2 ± 0.5 ka and 7.6 ± 0.6 ka. In contrast, the southern and eastern transects indicate that the LD persisted until 6.7 ± 0.4 ka. Based on these contrasts in timing of ice-margin retreat, we suggest that the LD separated into two smaller ice domes by 7.6 ± 0.6 ka, with one over central Quebec and Labrador and another over the Ungava Peninsula in northern Quebec (Fig. 1). Such rapid retreat could have been accelerated if a low ice saddle existed between the two ice domes prior to separation leading to enhanced ablation across lower elevations (Carlson et al., 2009a; Gregoire et al., 2012). Final deglaciation of these remnant ice domes occurred by 6.7 ± 0.4 ka according to our new \(^{10}\)Be chronology and a minimum-limiting \(^{14}\)C date (Guyard et al., 2011) (Fig. 6C, D, F). In contrast, minimum-limiting \(^{14}\)C dates from the Northern Hudson Bay region suggest that the remnant Keelewin and Foxe Domes (Fig. 1) had already disappeared by ~8 ka (Ross et al., 2012; Simon et al., 2014) (Fig. 6F).

Given evidence for a minimal role by increased dynamic ice flow in causing the accelerated retreat of the western LD margin (Carlson et al., 2009a; Ullman et al., 2015a; Stokes et al., 2016), an additional forcing must have acted with the overall enhanced boreal summer insolation and interglacial greenhouse gas forcings to cause a greater decrease in surface mass balance over this sector of the LD as compared to the more eastern LD sectors. We hypothesize that the transition from permanent to seasonal ice cover over Hudson Bay after its deglaciation at ~8.2 ka, and the associated changes in surface albedo and thermal capacity, provided the additional source of heat that accelerated melting of the western LD.

We test this hypothesis by replacing perennial ice cover over Hudson Bay with seasonally open water in our regional climate model simulations. This transition to seasonally open water conditions results in a warming of the June through October 2-m air temperature by > 2 °C (Fig. 5C). The change in surface water temperature creates lower surface pressure (Fig. 5D), which alters atmospheric circulation at the surface and aloft as anomalous anticyclonic flow that drives an enhancement of katabatic winds. The altered circulation advects anomalous heat from Hudson Bay into the atmosphere, producing warmer air temperatures over a substantial area around Hudson Bay and the LIS (Fig. 5C, E). Warming during the ablation season (June through October) enhances melt and surface runoff (Fig. 5F), resulting in the annual surface mass balance becoming more negative by 37% (Fig. 5G). The largest increase in ablation is over the low-sloping LIS margins, particularly along the western LD where the strongest heating occurs, consistent with our \(^{10}\)Be reconstructions of LD-margin retreat. As there is little change in annual precipitation over the LIS (Fig. 5H), changes in the surface mass balance are dominated by enhanced surface ablation. Our simulations suggest one potentially strong positive feedback that may have contributed to the deglaciation of Hudson Bay and rapid western LIS margin retreat, ultimately leading to full LIS deglaciation. This feedback warrants further exploration with a fully coupled regional atmosphere-ocean model capable of simulating the dynamics of freshwater runoff into Hudson Bay and the North Atlantic.
4.3. Laurentide ice-sheet contribution to Holocene sea-level rise

We use our new 10Be chronology for the LD to estimate its contributions to GMSL rise after the opening of Hudson Bay. At ~8.2 ka just following the opening of Hudson Bay, we estimate a LD volume of 3.6 ± 0.4 m of GMSL; we note that this is the total LIS volume at this time as all other domes were largely deglaciated (Fig. 6F) (Dykze, 2004; Ross et al., 2012; Simon et al., 2014). After separation of the LD into two domes by 7.6 ± 0.6 ka, these two domes have a summed volume between 1.0 ± 0.1 m and 0.2 ± 0.02 m GMSL depending on our maximum and minimum reconstructions (Fig. 1). The change in LD volume from ~8.2 ka to ~7.6 ka indicates 2.0–3.9 m of GMSL from the separation of the LD into two separate domes, which would explain most of the ~4.5 m GMSL rise from 8.2 to 7.6 ka (Fig. 6G) (Lambek et al., 2014). Our new LD 10Be chronology also precisely dates the final demise of the LIS and its contribution to GMSL rise at 6.7 ± 0.4 ka, which is concurrent with a slowing in the rate of GMSL rise (Fig. 6H) (Lambek et al., 2014), but is 1.5–3.0 ka earlier than LIS deglaciation in LIS models (Peltier, 2004; Bintanja and van de Wal, 2008; Tarasov et al., 2012; Abe-Ouchi et al., 2013).

GMSL was 3.4–6.3 m below present at 7.1–6.3 ka (Fig. 6H) (Lambek et al., 2014), indicating continued retreat of land-based ice elsewhere on the globe following LIS deglaciation. The Greenland Ice Sheet reached a minimum volume at ~4 ka and then accumulated the equivalent of ~0.2 m of GMSL in the late Holocene (Lecaivaller et al., 2014), and so cannot explain the 3.4–6.3 m rise in GMSL. This leaves the Antarctic ice sheets as the only remaining source that could have contributed a substantial amount to GMSL rise. Our inference is consistent with recent modeling results by Bradley et al. (2016) of far-field relative sea-level records that show ~5.8 m of GMSL rise sourced from Antarctic ice sheets since ~7 ka. In contrast, other Antarctic ice-sheet reconstructions range from <1 m to 2–3 m over the middle to late Holocene (Peltier, 2004; Whitehouse et al., 2012; Ivins et al., 2013; Briggs et al., 2014), which are inconsistent with our reconstruction and Bradley et al. (2016). Indeed, the breakup of the Ross Ice Shelf at ~5 ka (Yokoyama et al., 2016) and continued thinning of ice in Marie Byrd Land (Stone et al., 2003) are consistent with substantial loss of West Antarctic ice-sheet volume during the late Holocene.

5. Conclusions

Our results provide important constraints on the response times of ice sheets to peak deglacial forcings and abrupt climate events. We show, for the first time, that the LD margin may have responded to abrupt climate events at ~10.3 ka, 9.3 ka and 8.2 ka. Likewise, we find that the dominant loss of LIS volume occurred by 7.6 ± 0.6 ka, with final deglaciation of the LIS at 6.7 ± 0.4 ka. LIS deglaciation lagged ~4 ka behind peak insolation and atmospheric CO2 forcings, which is a shorter lag than exists in current LIS model reconstructions (Bintanja and van de Wal, 2008; Tarasov et al., 2012; Abe-Ouchi et al., 2013). In contrast, Antarctic ice sheets may have had a much longer response time to peak interglacial atmospheric CO2 and attendant global warming, with deglaciation continuing into the late Holocene. This finding has direct implications for how future simulations of Antarctic ice sheets are initiated (e.g., Golledge et al., 2015; Clark et al., 2016; DeConto and Pollard, 2016) as current models assume minimal Antarctic ice-volume change since the middle Holocene (Peltier, 2004; Whitehouse et al., 2012; Ivins et al., 2013; Briggs et al., 2014). Furthermore, recent model simulations of marine instability of the West Antarctic Ice Sheet predict rates of sea level rise of 1–2 cm a⁻¹ by 2100 C.E. (DeConto and Pollard, 2016). Our chronology shows that a land-based ice sheet of comparable size to the West Antarctic Ice Sheet may have retreated at rates that raised mid-Holocene GMSL at ~1 cm a⁻¹.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.quascirev.2016.09.014.

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